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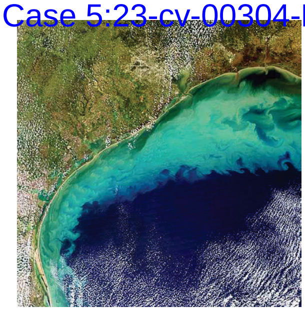

13

Ocean Acidification and Other Ocean Changes

KEY FINDINGS

1. The world's oceans have absorbed about 93% of the excess heat caused by greenhouse gas warming since the mid-20th century, making them warmer and altering global and regional climate feedbacks. Ocean heat content has increased at all depths since the 1960s and surface waters have warmed by about 1.3° ± 0.1°F (0.7° ± 0.08°C) per century globally since 1900 to 2016. Under a higher scenario, a global increase in average sea surface temperature of 4.9° ± 1.3°F (2.7° ± 0.7°C) by 2100 is projected, with even higher changes in some U.S. coastal regions. (*Very high confidence*)

2. The potential slowing of the Atlantic meridional overturning circulation (AMOC; of which the Gulf Stream is one component)—as a result of increasing ocean heat content and freshwater driven buoyancy changes—could have dramatic climate feedbacks as the ocean absorbs less heat and CO_2 from the atmosphere. This slowing would also affect the climates of North America and Europe. Any slowing documented to date cannot be directly tied to anthropogenic forcing primarily due to lack of adequate observational data and to challenges in modeling ocean circulation changes. Under a higher scenario (RCP8.5) in CMIP5 simulations, the AMOC weakens over the 21st century by 12% to 54% (*low confidence*).

3. The world's oceans are currently absorbing more than a quarter of the CO_2 emitted to the atmosphere annually from human activities, making them more acidic (*very high confidence*), with potential detrimental impacts to marine ecosystems. In particular, higher-latitude systems typically have a lower buffering capacity against pH change, exhibiting seasonally corrosive conditions sooner than low-latitude systems. Acidification is regionally increasing along U.S. coastal systems as a result of upwelling (for example, in the Pacific Northwest) (*high confidence*), changes in freshwater inputs (for example, in the Gulf of Maine) (*medium confidence*), and nutrient input (for example, in agricultural watersheds and urbanized estuaries) (*high confidence*). The rate of acidification is unparalleled in at least the past 66 million years (*medium confidence*). Under the higher scenario (RCP8.5), the global average surface ocean acidity is projected to increase by 100% to 150% (*high confidence*).

4. Increasing sea surface temperatures, rising sea levels, and changing patterns of precipitation, winds, nutrients, and ocean circulation are contributing to overall declining oxygen concentrations at intermediate depths in various ocean locations and in many coastal areas. Over the last half century, major oxygen losses have occurred in inland seas, estuaries, and in the coastal and open ocean (*high confidence*). Ocean oxygen levels are projected to decrease by as much as 3.5% under the higher scenario (RCP8.5) by 2100 relative to preindustrial values (*high confidence*).

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13.0 A Changing Ocean

Anthropogenic perturbations to the global Earth system have included important alterations in the chemical composition, temperature, and circulation of the oceans. Some of these changes will be distinguishable from the background natural variability in nearly half of the global open ocean within a decade, with important consequences for marine ecosystems and their services.[1] However, the timeframe for detection will vary depending on the parameter featured.[2, 3]

13.1 Ocean Warming

13.1.1 General Background

Approximately 93% of excess heat energy trapped since the 1970s has been absorbed into the oceans, lessening atmospheric warming and leading to a variety of changes in ocean conditions, including sea level rise and ocean circulation (see Ch. 2: Physical Drivers of Climate Change, Ch. 6: Temperature Change, and Ch. 12: Sea Level Rise in this report).[1, 4] This is the result of the high heat capacity of seawater relative to the atmosphere, the relative area of the ocean compared to the land, and the ocean circulation that enables the transport of heat into deep waters. This large heat absorption by the oceans moderates the effects of increased anthropogenic greenhouse emissions on terrestrial climates while altering the fundamental physical properties of the ocean and indirectly impacting chemical properties such as the biological pump through increased stratification.[1, 5] Although upper ocean temperature varies over short- and medium timescales (for example, seasonal and regional patterns), there are clear long-term increases in surface temperature and ocean heat content over the past 65 years.[4, 6, 7]

13.1.2 Ocean Heat Content

Ocean heat content (OHC) is an ideal variable to monitor changing climate as it is calculated using the entire water column, so ocean warming can be documented and compared between particular regions, ocean basins, and depths. However, for years prior to the 1970s, estimates of ocean uptake are confined to the upper ocean (up to 700 m) due to sparse spatial and temporal coverage and limited vertical capabilities of many of the instruments in use. OHC estimates are improved for time periods after 1970 with increased sampling coverage and depth.[4, 8] Estimates of OHC have been calculated going back to the 1950s using averages over longer time intervals (i.e., decadal or 5-year intervals) to compensate for sparse data distributions, allowing for clear long-term trends to emerge (e.g., Levitus et al. 2012[7]).

From 1960 to 2015, OHC significantly increased for both 0–700 and 700–2,000 m depths, for a total ocean warming of about $33.5 \pm 7.0 \times 10^{22}$ J (a net heating of 0.37 ± 0.08 W/m^2; Figure 13.1).[6] During this period, there is evidence of an acceleration of ocean warming beginning in 1998,[9] with a total heat increase of about 15.2×10^{22} J.[6] Robust ocean warming occurs in the upper 700 m and is slow to penetrate into the deep ocean. However, the 700–2,000 m depths constitute an increasing portion of the total ocean energy budget as compared to the surface ocean (Figure 13.1).[6] The role of the deep ocean (below 2,000 m [6,600 ft]) in ocean heat uptake remains uncertain, both in the magnitude but also the sign of the uptake.[10, 11] Penetration of surface waters to the deep ocean is a slow process, which means that while it takes only about a decade for near-surface temperatures to respond to increased heat energy, the deep ocean will continue to warm, and as a result sea levels will rise for centuries to millennia even if all further emissions cease.[4]

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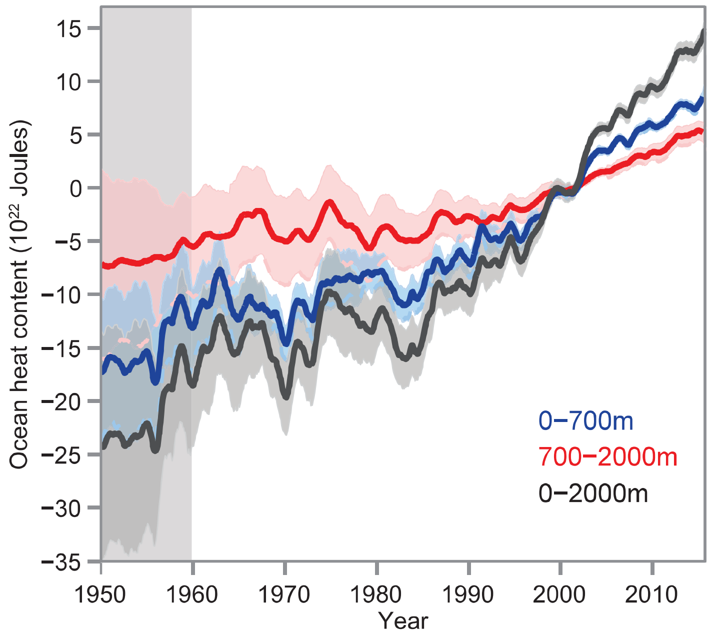

Figure 13.1: Global Ocean heat content change time series. Ocean heat content from 0 to 700 m (blue), 700 to 2,000 m (red), and 0 to 2,000 m (dark gray) from 1955 to 2015 with an uncertainty interval of ±2 standard deviations shown in shading. All time series of the analysis performed by Cheng et al.[6] are smoothed by a 12-month running mean filter, relative to the 1997–2005 base period. (Figure source: Cheng et al. 2017[6]).

Several sources have documented warming in all ocean basins from 0–2,000 m depths over the past 50 years (Figure 13.2).[6, 7, 12] Annual fluctuations in surface temperatures and OHC are attributed to the combination of a long-term secular trend and decadal and smaller time scale variations, such as the Pacific Decadal Oscillation (PDO) and the Atlantic Multidecadal Oscillation (AMO) (Ch. 5: Circulation & Variability; Ch. 12: Sea Level Rise).[13, 14] The transport of heat to the deep ocean is likely linked to the strength of the Atlantic Meridional Overturning Circulation (see Section 13.2.1), where the Atlantic and Southern Ocean accounts for the dominant portion of total OHC change at the 700–2,000 m depth.[6, 8, 9, 15] Decadal variability in ocean heat uptake is mostly attributed to ENSO phases (with El Niños warming and La Niñas cooling). For instance, La Niña conditions over the past decade have led to colder ocean temperatures in the eastern tropical Pacific.[6, 8, 9, 16] For the Pacific and Indian Oceans, the decadal shifts are primarily observed in the upper 350 m depth, likely due to shallow subtropical circulation, leading to an abrupt increase of OHC in the Indian Ocean carried by the Indonesian throughflow from the Pacific Ocean over the last decade.[9] Although there is natural variability in ocean temperature, there remain clear increasing trends due to anthropogenic influences.

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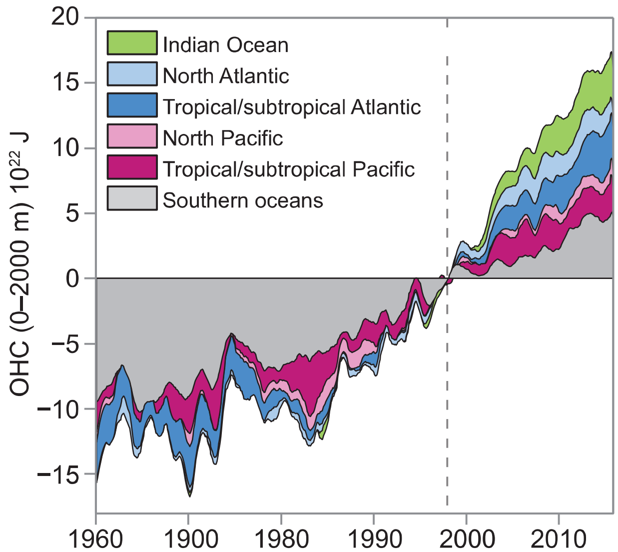

Figure 13.2: Ocean heat content changes from 1960 to 2015 for different ocean basins for 0 to 2,000 m depths. Time series is relative to the 1997–1999 base period and smoothed by a 12-month running filter by Cheng et al.[6] The curves are additive, and the ocean heat content changes in different ocean basins are shaded in different colors (Figure source: Cheng et al. 2017[6]).

13.1.3 Sea Surface Temperature and U.S. Regional Warming

In addition to OHC, sea surface temperature (SST) measurements are widely available. SST measurements are useful because 1) the measurements have been taken over 150 years (albeit using different platforms, instruments, and depths through time); 2) SST reflects the lower boundary condition of the atmosphere; and 3) SST can be used to predict specific regional impacts of global warming on terrestrial and coastal systems.[15, 17, 18] Globally, surface ocean temperatures have increased by 1.3° ± 0.1°F (0.70° ± 0.08°C) per century from 1900 to 2016 for the Extended Reconstructed Sea Surface Temperature version 4 (ERSST v4) record.[19] All U.S. coastal waters have warmed by more than 0.7°F (0.4°C) over this period as shown in both Table 13.1 and Chapter 6: Temperature Change, Figure 6.6. During the past 60 years, the rates of increase of SSTs for the coastal waters of three U.S. regions were above the global average rate. These included the waters around Alaska, the Northeast, and the Southwest (Table 13.1). Over the last decade, some regions have experienced increased high ocean temperature anomalies. SST in the Northeast has warmed faster than 99% of the global ocean since 2004, and a peak temperature for the region in 2012 was part of a large "ocean heat wave" in the Northwest Atlantic that persisted for nearly 18 months.[20, 21] Projections indicate that the Northeast will continue to warm more quickly than other ocean regions through the end of the century.[22] In the Northwest, a resilient ridge of high pressure over the North American West Coast suppressed storm activity and mixing, which intensified heat in the upper ocean in a phenomenon known as "The Blob".[23] Anomalously warm waters persisted in the coastal waters of the Alaskan and Pacific Northwest from 2013 until 2015. Under a higher scenario (RCP8.5), SSTs are projected to increase by an additional 4.9°F (2.7°C) by 2100 (Figure 13.3), whereas for a lower scenario (RCP4.5) the SST increase would be 2.3°F (1.3°C).[24] In all U.S. coastal regions, the warming since 1901 is detectable compared to natural variability and attributable to anthropogenic forcing, according to an analysis of the CMIP5 models (Ch. 6: Temperature Change, Figure 6.5).

Table 13.1. Historical sea surface temperature trends (°C per century) and projected trends by 2080 (°C) for eight U.S. coastal regions and globally. Historical temperature trends are presented for the 1900–2016 and 1950–2016 periods with 95% confidence level, observed using the Extended Reconstructed Sea Surface Temperature version 4 (ERSSTv4).[19] Global and regional predictions are calculated for lower and higher scenarios (RCP4.5 and RCP8.5, respectively) with 80% spread of all the CMIP5 members compared to the 1976–2005 period.[151] The historical trends were analyzed for the latitude and longitude in the table, while the projected trends were analyzed for the California Current instead of the Northwest and Southwest separately and for the Bering Sea in Alaska (NOAA).

<table>
<tr><th rowspan="2">Region</th><th rowspan="2">Latitude and Longitude</th><th colspan="2">Historical Trend (°C/100 years)</th><th colspan="2">Projected Trend 2080 (relative to 1976–2005 climate) (°C)</th></tr>
<tr><th>1900–2016</th><th>1950–2016</th><th>RCP4.5</th><th>RCP8.5</th></tr>
<tr><td>Global</td><td></td><td>0.70 ± 0.08</td><td>1.00 ± 0.11</td><td>1.3 ± 0.6</td><td>2.7 ± 0.7</td></tr>
<tr><td>Alaska</td><td>50°–66°N, 150°–170°W</td><td>0.82 ± 0.26</td><td>1.22 ± 0.59</td><td>2.5 ± 0.6</td><td>3.7 ± 1.0</td></tr>
<tr><td>Northwest (NW)</td><td>40°–50°N, 120°–132°W</td><td>0.64 ± 0.30</td><td>0.68 ± 0.70</td><td rowspan="2">1.7 ± 0.4</td><td rowspan="2">2.8 ± 0.6</td></tr>
<tr><td>Southwest (SW)</td><td>30°–40°N, 116°–126°W</td><td>0.73 ± 0.33</td><td>1.02 ± 0.79</td></tr>
<tr><td>Hawaii (HI)</td><td>18°–24°N, 152°–162°W</td><td>0.58 ± 0.19</td><td>0.46 ± 0.39</td><td>1.6 ± 0.4</td><td>2.8 ± 0.6</td></tr>
<tr><td>Northeast (NE)</td><td>36°–46°N, 64°–76°W</td><td>0.63 ± 0.31</td><td>1.10 ± 0.71</td><td>2.0 ± 0.3</td><td>3.2 ± 0.6</td></tr>
<tr><td>Southeast (SE)</td><td>24°–34°N, 64°–80°W</td><td>0.40 ± 0.18</td><td>0.13 ± 0.34</td><td>1.6 ± 0.3</td><td>2.7 ± 0.4</td></tr>
<tr><td>Gulf of Mexico (GOM)</td><td>20°–30°N, 80°–96°W</td><td>0.52 ± 0.14</td><td>0.37 ± 0.27</td><td>1.6 ± 0.3</td><td>2.8 ± 0.3</td></tr>
<tr><td>Caribbean</td><td>10°–20°N, 66°–86°W</td><td>0.76 ± 0.15</td><td>0.77 ± 0.32</td><td>1.5 ± 0.4</td><td>2.6 ± 0.3</td></tr>
</table>

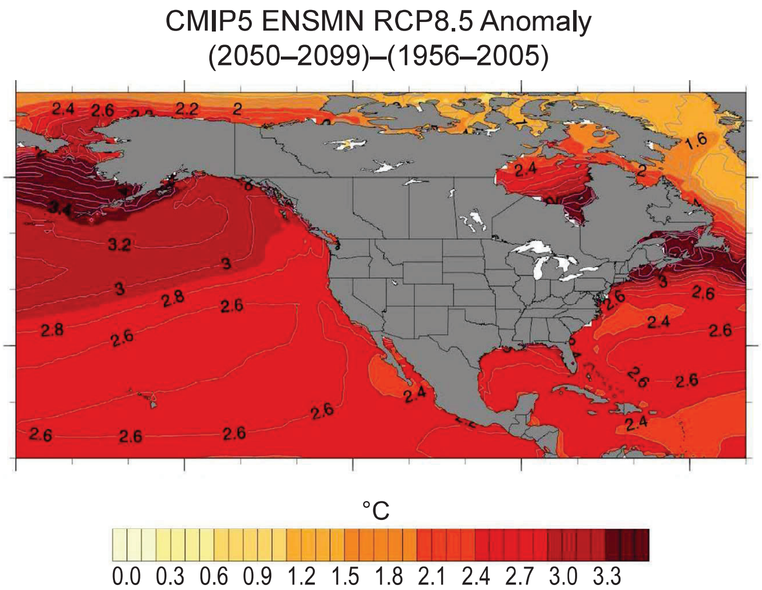

Figure 13.3: Projected changes in sea surface temperature (°C) for the coastal United States under the higher scenario (RCP8.5). Projected anomalies for the 2050–2099 period are calculated using a comparison from the average sea surface temperatures over 1956–2005. Projected changes are examined using the Coupled Model Intercomparison Project Phase 5 (CMIP5) suite of model simulations. (Figure source: NOAA).

13.1.4 Ocean Heat Feedback

The residual heat not taken up by the oceans increases land surface temperatures (approximately 3%) and atmospheric temperatures (approximately 1%), and melts both land and sea ice (approximately 3%), leading to sea level rise (see Ch. 12: Sea Level Rise).[4, 6, 25] The meltwater from land and sea ice amplifies further subsurface ocean warming and ice shelf melting, primarily due to increased thermal stratification, which reduces the ocean's efficiency in transporting heat to deep waters.[4] Surface ocean stratification has increased by about 4% during the period 1971–2010[26] due to thermal heating and freshening from increased freshwater inputs (precipitation and evaporation changes and land and sea ice melting). The increase of ocean stratification will contribute to further feedback of ocean warming and, indirectly, mean sea level. In addition, increases in stratification are associated with suppression of tropical cyclone intensification,[27] retreat of the polar ice sheets,[28] and reductions of the convective mixing at higher latitudes that transports heat to the deep ocean through the Atlantic Meridional Overturning Circulation.[29] Ocean heat uptake therefore represents an important feedback that will have a significant influence on future shifts in climate (see Ch. 2: Physical Drivers of Climate Change).

13.2 Ocean Circulation

13.2.1 Atlantic Meridional Overturning Circulation

The Atlantic Meridional Overturning Circulation (AMOC) refers to the three-dimensional, time-dependent circulation of the Atlantic Ocean, which has been a high priority topic of study in recent decades. The AMOC plays an important role in climate through its transport of heat, freshwater, and carbon (e.g., Johns et al. 2011;[30] McDonagh et al. 2015;[31] Talley et al. 2016[32]). AMOC-associated poleward heat transport substantially contributes to North American and continental European climate (see Ch. 5: Circulation and Variability). The Gulf Stream, in contrast to other western boundary currents, is expected to slow down because of the weakening of the AMOC, which would impact the Euro-

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pean climate.[33] Variability in the AMOC has been attributed to wind forcing on intra-annual time scales and to geostrophic forces on interannual to decadal timescales.[34] Increased freshwater fluxes from melting Arctic Sea and land ice can weaken open ocean convection and deep-water formation in the Labrador and Irminger Seas, which could weaken the AMOC (Ch. 11: Arctic Changes; also see Ch. 5, Section 5.2.3: North Atlantic Oscillation and Northern Annular Mode).[29, 33]

While one recent study has suggested that the AMOC has slowed since preindustrial times[29] and another suggested slowing on faster time scales,[35] there is at present insufficient observational evidence to support a finding of long term slowdown of AMOC strength over the 20th century[4] or within the last 50 years[34] as decadal ocean variability can obscure long-term trends. Some studies show long-term trends,[36, 37] but the combination of sparse data and large seasonal variability may also lead to incorrect interpretations (e.g., Kanzow et al. 2010[38]). Several recent high resolution modeling studies constrained with the limited existing observational data[39] and/or with reconstructed freshwater fluxes[40] suggest that the recently observed AMOC slowdown at 26°N (off the Florida coast) since 2004 (e.g., as described in Smeed et al. 2014[35]) is mainly due to natural variability, and that anthropogenic forcing has not yet caused a significant AMOC slowdown. In addition, direct observations of the AMOC in the South Atlantic fail to unambiguously demonstrate anthropogenic trends (e.g., Dong et al. 2015;[41] Garzoli et al. 2013[42]).

Under a higher scenario (RCP8.5) in CMIP5 simulations, it is very likely that the AMOC will weaken over the 21st century. The projected decline ranges from 12% to 54%,[43] with the range width reflecting substantial uncertainty in quantitative projections of AMOC behavior. In lower scenarios (like RCP4.5), CMIP5 models predict a 20% weakening of the AMOC during the first half of the 21st century and a stabilization and slight recovery after that.[44] The projected slowdown of the AMOC will be counteracted by the warming of the deep ocean (below 700 m [2,300 ft]), which will tend to strengthen the AMOC.[45] The situation is further complicated due to the known bias in coupled climate models related to the direction of the salinity transport in models versus observations, which is an indicator of AMOC stability (e.g., Drijhout et al. 2011;[46] Bryden et al. 2011;[47] Garzoli et al. 2013[42]). Some argue that coupled climate models should be corrected for this known bias and that AMOC variations could be even larger than the gradual decrease most models predict if the AMOC were to shut down completely and "flip states".[48] Any AMOC slowdown could result in less heat and CO_2 absorbed by the ocean from the atmosphere, which is a positive feedback to climate change (also see Ch. 2: Physical Drivers of Climate Change).[49, 50, 51]

13.2.2 Changes in Salinity Structure

As a response to warming, increased atmospheric moisture leads to stronger evaporation or precipitation in terrestrial and oceanic environments and melting of land and sea ice. Approximately 80% of precipitation/evaporation events occur over the ocean, leading to patterns of higher salt content or freshwater anomalies and changes in ocean circulation (see Ch. 2: Physical Drivers of Climate Change and Ch. 6: Temperature Change).[52] Over 1950–2010, average global amplification of the surface salinity pattern amounted to 5.3%; where fresh regions in the ocean became fresher and salty regions became saltier.[53] However, the long-term trends of these physical and chemical changes to the ocean are difficult to isolate from natural large-scale variability. In particular, ENSO displays particular salinity and precipitation/evaporation patterns that skew the trends. More research and data are neces-

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sary to better model changes to ocean salinity. Several models have shown a similar spatial structure of surface salinity changes, including general salinity increases in the subtropical gyres, a strong basin-wide salinity increase in the Atlantic Ocean, and reduced salinity in the western Pacific warm pools and the North Pacific subpolar regions.[52, 53] There is also a stronger distinction between the upper salty thermocline and fresh intermediate depth through the century. The regional changes in salinity to ocean basins will have an overall impact on ocean circulation and net primary production, leading to corresponding carbon export (see Ch. 2: Physical Drivers of Climate Change). In particular, the freshening of the Arctic Ocean due to melting of land and sea ice can lead to buoyancy changes which could slow down the AMOC (see Section 13.2.1).

13.2.3 Changes in Upwelling

Significant changes to ocean stratification and circulation can also be observed regionally, along the eastern ocean boundaries and at the equator. In these areas, wind-driven upwelling brings colder, nutrient- and carbon-rich water to the surface; this upwelled water is more efficient in heat and anthropogenic CO_2 uptake. There is some evidence that coastal upwelling in mid- to high-latitude eastern boundary regions has increased in intensity and/or frequency,[54] but in more tropical areas of the western Atlantic, such as in the Caribbean Sea, it has decreased between 1990 and 2010.[55, 56] This has led to a decrease in primary productivity in the southern Caribbean Sea.[55] Within the continental United States, the California Current is experiencing fewer (by about 23%–40%) but stronger upwelling events.[57, 58, 59] Stronger offshore upwelling combined with cross-shelf advection brings nutrients from the deeper ocean but also increased offshore transport.[60] The net nutrient load in the coastal regions is responsible for increased productivity and ecosystem function.

IPCC 2013 concluded that there is low confidence in the current understanding of how eastern upwelling systems will be altered under future climate change because of the obscuring role of multidecadal climate variability.[26] However, subsequent studies show that by 2100, upwelling is predicted to start earlier in the year, end later, and intensify in three of the four major eastern boundary upwelling systems (not in the California Current).[61] In the California Current, upwelling is projected to intensify in spring but weaken in summer, with changes emerging from the envelope of natural variability primarily in the second half of the 21st century.[62] Southern Ocean upwelling will intensify while the Atlantic equatorial upwelling systems will weaken.[57, 61] The intensification is attributed to the strengthening of regional coastal winds as observations already show,[58] and model projections under the higher scenario (RCP8.5) estimate wind intensifying near poleward boundaries (including northern California Current) and weakening near equatorward boundaries (including southern California Current) for the 21st century.[61, 63]

13.3 Ocean Acidification

13.3.1 General Background

In addition to causing changes in climate, increasing atmospheric levels of carbon dioxide (CO_2) from the burning of fossil fuels and other human activities, including changes in land use, have a direct effect on ocean carbonate chemistry that is termed ocean acidification.[64, 65] Surface ocean waters absorb part of the increasing CO_2 in the atmosphere, which causes a variety of chemical changes in seawater: an increase in the partial pressure of CO_2 ($pCO_{2,sw}$), dissolved inorganic carbon (DIC), and the concentration of hydrogen and bicarbonate ions and a decrease in the concentration of carbonate ions (Figure 13.4). In brief, CO_2 is an acid gas that combines with water to form carbonic acid, which then dissociates

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to hydrogen and bicarbonate ions. Increasing concentrations of seawater hydrogen ions result in a decrease of carbonate ions through their conversion to bicarbonate ions. The concentration of carbonate ions in seawater affects saturation states for calcium carbonate compounds, which many marine species use to build their shells and skeletons. Ocean acidity refers to the concentration of hydrogen ions in ocean seawater regardless of ocean pH, which is fundamentally basic (e.g., pH > 7). Ocean surface waters have become 30% more acidic over the last 150 years as they have absorbed large amounts of CO_2 from the atmosphere,[66] and anthropogenically sourced CO_2 is gradually invading into oceanic deep waters. Since the preindustrial period, the oceans have absorbed approximately 29% of all CO_2 emitted to the atmosphere.[67] Oceans currently absorb about 26% of the human-caused CO_2 anthropogenically emitted into the atmosphere.[67]

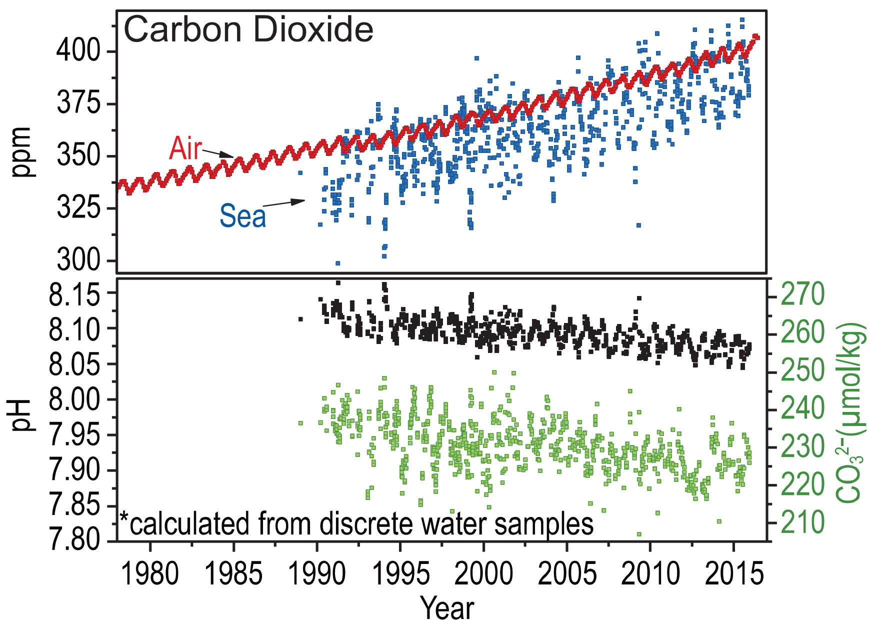

Figure 13.4: Trends in surface (< 50 m) ocean carbonate chemistry calculated from observations obtained at the Hawai'i Ocean Time-series (HOT) Program in the North Pacific over 1988–2015. The upper panel shows the linked increase in atmospheric (red points) and seawater (blue points) CO_2 concentrations. The bottom panel shows a decline in seawater pH (black points, primary y-axis) and carbonate ion concentration (green points, secondary y-axis). Ocean chemistry data were obtained from the Hawai'i Ocean Time-series Data Organization & Graphical System (HOT-DOGS, http://hahana.soest.hawaii.edu/hot/hot-dogs/index.html). (Figure source: NOAA).

13.3.2 Open Ocean Acidification

Surface waters in the open ocean experience changes in carbonate chemistry reflective of large-scale physical oceanic processes (see Ch. 2: Physical Drivers of Climate Change). These processes include both the global uptake of atmospheric CO_2 and the shoaling of naturally acidified subsurface waters due to vertical mixing and upwelling. In general, the rate of ocean acidification in open ocean surface waters at a decadal time-scale closely approximates the rate of atmospheric CO_2 increase.[68] Large, multidecadal phenomena such as the Atlantic Multidecadal Oscillation and Pacific Decadal Oscillation can add variability to the observed rate of change.[68]

13.3.3 Coastal Acidification

Coastal shelf and nearshore waters are influenced by the same processes as open ocean surface waters such as absorption of atmospheric CO_2 and upwelling, as well as a number of additional, local-level processes, including freshwater, nutrient, sulfur, and nitrogen inputs.[69, 70] Coastal acidification generally exhibits higher-frequency variability and short-term episodic events relative to open-ocean acidification.[71, 72, 73, 74] Upwelling is of particular importance in coastal waters, especially along the U.S. West Coast. Deep waters that shoal with upwelling are enriched in CO_2 due to uptake of anthropogenic atmospheric CO_2 when last in contact with the atmosphere, coupled with deep water respiration processes and lack of gas exchange with the atmosphere.[65, 75] Freshwater inputs to coastal waters change seawater chemistry in ways that make it more susceptible to acidification, largely by freshening ocean waters and contributing varying amounts of dissolved inorganic carbon (DIC), total alkalinity (TA), dissolved and particulate organic carbon, and nutrients from riverine and estuarine sources. Coastal waters of the East Coast and mid-Atlantic are far more influenced by freshwater inputs than are Pacific Coast waters.[76] Coastal waters can episodically experience riverine and glacial melt plumes that create conditions in which seawater can dissolve calcium carbonate structures.[77, 78] While these processes have persisted historically, climate-induced increases in glacial melt and high-intensity precipitation events can yield larger freshwater plumes than have occurred in the past. Nutrient runoff can increase coastal acidification by creating conditions that enhance biological respiration. In brief, nutrient loading typically promotes phytoplankton blooms, which, when they die, are consumed by bacteria. Bacteria respire CO_2 and thus bacterial blooms can result in acidification events whose intensity depends on local hydrographic conditions, including water column stratification and residence time.[72] Long-term changes in nutrient loading, precipitation, and/or ice melt may also impart long-term, secular changes in the magnitude of coastal acidification.

13.3.4 Latitudinal Variation

Ocean carbon chemistry is highly influenced by water temperature, largely because the solubility of CO_2 in seawater increases as water temperature declines. Thus, cold, high-latitude surface waters can retain more CO_2 than warm, lower-latitude surface waters.[76, 79] Because carbonate minerals also more readily dissolve in colder waters, these waters can more regularly become undersaturated with respect to calcium carbonate whereby mineral dissolution is energetically favored. This chemical state, often referred to as seawater being "corrosive" to calcium carbonate, is important when considering the ecological implications of ocean acidification as many species make structures such as shells and skeletons from calcium carbonate. Seawater conditions undersaturated with respect to calcium carbonate are common at depth, but currently and historically rare at the surface and near-surface.[80] Some high-latitude surface

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and near-surface waters now experience such corrosive conditions, which are rarely documented in low-latitude surface or near-surface systems. For example, corrosive conditions at a range of ocean depths have been documented in the Arctic and northeastern Pacific Oceans.[74, 79, 81, 82] Storm-induced upwelling could cause undersaturation in tropical areas in the future.[83] It is important to note that low-latitude waters are experiencing a greater absolute rate of change in calcium carbonate saturation state than higher latitudes, though these low-latitude waters are not approaching the undersaturated state except within nearshore or some benthic habitats.[84]

13.3.5 Paleo Evidence

Evidence suggests that the current rate of ocean acidification is the fastest in the last 66 million years (the K-Pg boundary) and possibly even the last 300 million years (when the first pelagic calcifiers evolved providing proxy information and also a strong carbonate buffer, characteristic of the modern ocean).[85, 86] The Paleo-Eocene Thermal Maximum (PETM; around 56 million years ago) is often referenced as the closest analogue to the present, although the overall rate of change in CO_2 conditions during that event (estimated between 0.6 and 1.1 GtC/year) was much lower than the current increase in atmospheric CO_2 of 10 GtC/year.[86, 87] The relatively slower rate of atmospheric CO_2 increase at the PETM likely led to relatively small changes in carbonate ion concentration in seawater compared with the contemporary acidification rate, due to the ability of rock weathering to buffer the change over the longer time period.[86] Some of the presumed acidification events in Earth's history have been linked to selective extinction events suggestive of how guilds of species may respond to the current acidification event.[85]

13.3.6 Projected Changes

Projections indicate that by the end of the century under higher scenarios, such as SRES A1FI or RCP8.5, open-ocean surface pH will decline from the current average level of 8.1 to a possible average of 7.8 (Figure 13.5).[1] When the entire ocean volume is considered under the same scenario, the volume of waters undersaturated with respect to calcium carbonate could expand from 76% in the 1990s to 91% in 2100, resulting in a shallowing of the saturation horizons—depths below which undersaturation occurs.[1, 88] Saturation horizons, which naturally vary among ocean basins, influence ocean carbon cycles and organisms with calcium carbonate structures, especially as they shoal into the zones where most biota lives.[81, 89] As discussed above, for a variety of reasons, not all ocean and coastal regions will experience acidification in the same way depending on other compounding factors. For instance, recent observational data from the Arctic Basin show that the Beaufort Sea became undersaturated, for part of the year, with respect to aragonite in 2001, while other continental shelf seas in the Arctic Basin are projected to do so closer to the middle of the century (e.g., the Chukchi Sea in about 2033 and Bering Sea in about 2062).[90] Deviation from the global average rate of acidification will be especially true in coastal and estuarine areas where the rate of acidification is influenced by other drivers than atmospheric CO_2, some of which are under the control of local management decisions (for example, nutrient pollution loads).

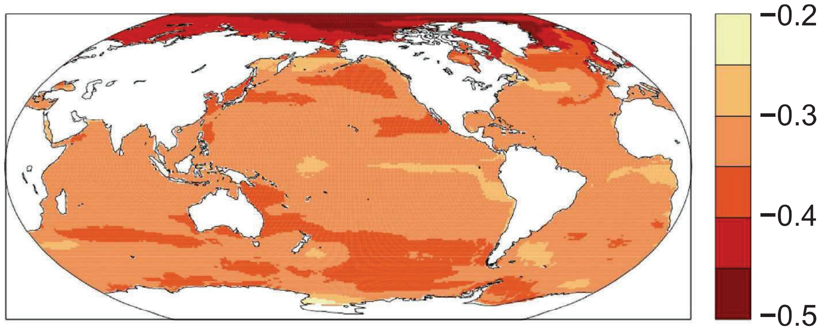

Figure 13.5: Predicted change in sea surface pH in 2090–2099 relative to 1990–1999 under the higher scenario (RCP8.5), based on the Community Earth System Models–Large Ensemble Experiments CMIP5 (Figure source: adapted from Bopp et al. 2013[24]).

13.4 Ocean Deoxygenation

13.4.1 General Background

Oxygen is essential to most life in the ocean, governing a host of biogeochemical and biological processes. Oxygen influences metabolic, physiological, reproductive, behavioral, and ecological processes, ultimately shaping the composition, diversity, abundance, and distribution of organisms from microbes to whales. Increasingly, climate-induced oxygen loss (deoxygenation) associated with ocean warming and reduced ventilation to deep waters has become evident locally, regionally, and globally. Deoxygenation can also be attributed to anthropogenic nutrient input, especially in the coastal regions, where the nutrients can lead to the proliferation of primary production and, consequently, enhanced drawdown of dissolved oxygen by microbes.[91] In addition, acidification (Section 13.2) can co-occur with deoxygenation as a result of warming-enhanced biological respiration.[92] As aerobic organisms respire, O_2 is consumed and CO_2 is produced. Understanding the combined effect of both low O_2 and low pH on marine ecosystems is an area of active research.[93] Warming also raises biological metabolic rates which, in combination with intensified coastal and estuarine stratification, exacerbates eutrophication-induced hypoxia. We now see earlier onset and longer periods of seasonal hypoxia in many eutrophic sites, most of which occur in areas that are also warming.[91]

13.4.2 Climate Drivers of Ocean Deoxygenation

Global ocean deoxygenation is a direct effect of warming. Ocean warming reduces the solubility of oxygen (that is, warmer water can hold less oxygen) and changes physical mixing (for example, upwelling and circulation) of oxygen in the oceans. The increased temperature of global oceans accounts for about 15% of current global oxygen loss,[94] although changes in temperature and oxygen are not uniform throughout the ocean.[15] Warming also exerts direct influence on thermal stratification and enhances salinity stratification through ice melt and climate change-associated precipitation effects. Intensified stratification leads to reduced ventilation (mixing of oxygen into the ocean interior) and accounts for up to 85% of global ocean oxygen loss.[94] Effects of ocean temperature change and stratification on oxygen loss are strongest in intermediate or mode waters at bathyal depths (in general, 200–3,000 m) and also nearshore and in the open ocean; these changes are especially evident in tropical and subtropical waters globally, in the Eastern Pacific,[95] and in the Southern Ocean.[94]

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There are also other, less direct effects of global temperature increase. Warming on land reduces terrestrial plant water efficiency (through effects on stomata; see Ch. 8: Drought, Floods, and Wildfires, Key Message 3), leading to greater runoff, on average, into coastal zones (see Ch. 8: Drought, Floods, and Wildfires for other hydrological effects of warming) and further enhancing hypoxia potential because greater runoff can mean more nutrient transport (See Ch. 2: Physical Drivers of Climate Change).[96, 97] Estuaries, especially ones with minimal tidal mixing, are particularly vulnerable to oxygen-depleted dead zones from the enhanced runoff and stratification. Warming can induce dissociation of frozen methane in gas hydrates buried on continental margins, leading to further drawdown of oxygen through aerobic methane oxidation in the water column.[98] On eastern ocean boundaries, warming can enhance the land–sea temperature differential, causing increased upwelling due to higher winds with (a) greater nutrient input leading to production, sinking, decay, and biochemical drawdown of oxygen and (b) upwelling of naturally low-oxygen, high-CO_2 waters onto the upper slope and shelf environments.[58, 65] However, in the California Current, upwelling intensification has occurred only in the poleward regions (north of San Francisco), and the drivers may not be associated with land–sea temperature differences.[63] Taken together, the effects of warming are manifested as low-oxygen water in open oceans are being transported to and upwelled along coastal regions. These low-oxygen upwelled waters are then coupled with eutrophication-induced hypoxia, further reducing oxygen content in coastal areas.

Changes in precipitation, winds, circulation, airborne nutrients, and sea level can also contribute to ocean deoxygenation. Projected increases in precipitation in some regions will intensify stratification, reducing vertical mixing and ventilation, and intensify nutrient input to coastal waters through excess runoff, which leads to increased algal biomass and concurrent dissolved oxygen consumption via community respiration.[99] Coastal wetlands that might remove these nutrients before they reach the ocean may be lost through rising sea level, further exacerbating hypoxia.[97] Some observations of oxygen decline are linked to regional changes in circulation involving low-oxygen water masses. Enhanced fluxes of airborne iron and nitrogen are interacting with natural climate variability and contributing to fertilization, enhanced respiration, and oxygen loss in the tropical Pacific.[100]

13.4.3 Biogeochemical Feedbacks of Deoxygenation to Climate and Elemental Cycles

Climate patterns and ocean circulation have a large effect on global nitrogen and oxygen cycles, which in turn affect phosphorus and trace metal availability and generate feedbacks to the atmosphere and oceanic production. Global ocean productivity may be affected by climate-driven changes below the tropical and subtropical thermocline which control the volume of suboxic waters (< 5 micromolar O_2), and consequently the loss of fixed nitrogen through denitrification.[101, 102] The extent of suboxia in the open ocean also regulates the production of the greenhouse gas nitrous oxide (N_2O); as oxygen declines, greater N_2O production may intensify global warming, as N_2O is about 310 times more effective at trapping heat than CO_2 (see Ch. 2: Physical Drivers of Climate Change, Section 2.3.2).[103, 104] Production of hydrogen sulfide (H_2S, which is highly toxic) and intensified phosphorus recycling can occur at low oxygen levels.[105] Other feedbacks may emerge as oxygen minimum zone (OMZ) shoaling diminishes the depths of diurnal vertical migrations by fish and invertebrates, and as their huge biomass and associated oxygen consumption deplete oxygen.[106]

13.4.4 Past Trends

Over hundreds of millions of years, oxygen has varied dramatically in the atmosphere and ocean and has been linked to biodiversity gains and losses.[107, 108] Variation in oxygenation in the paleo record is very sensitive to climate—with clear links to temperature and often CO_2 variation.[109] OMZs expand and contract in synchrony with warming and cooling events, respectively.[110] Episodic climate events that involve rapid temperature increases over decades, followed by a cool period lasting a few hundred years, lead to major fluctuations in the intensity of Pacific and Indian Ocean OMZs (i.e., DO of < 20 μM). These events are associated with rapid variations in North Atlantic deep water formation.[111] Ocean oxygen fluctuates on glacial-interglacial timescales of thousands of years in the Eastern Pacific.[112, 113]

13.4.5 Modern Observations (last 50+ years)

Long-term oxygen records made over the last 50 years reflect oxygen declines in inland seas,[114, 115, 116] in estuaries,[117, 118] and in coastal waters.[119, 120, 121, 122] The number of coastal, eutrophication-induced hypoxic sites in the United States has grown dramatically over the past 40 years.[123] Over larger scales, global syntheses show hypoxic waters have expanded by 4.5 million km^2 at a depth of 200 m,[95] with widespread loss of oxygen in the Southern Ocean,[94] Western Pacific,[124] and North Atlantic.[125] Overall oxygen declines have been greater in coastal oceans than in the open ocean[126] and often greater inshore than offshore.[127] The emergence of a deoxygenation signal in regions with naturally high oxygen variability will unfold over longer time periods (20–50 years from now).[128]

13.4.6 Projected Changes

Global Models

Global models generally agree that ocean deoxygenation is occurring; this finding is also reflected in in situ observations from past 50 years. Compilations of 10 Earth System models predict a global average loss of oxygen of −3.5% (higher scenario, RCP8.5) to −2.4% (lower scenario, RCP4.5) by 2100, but much stronger losses regionally, and in intermediate and mode waters (Figure 13.6).[24] The North Pacific, North Atlantic, Southern Ocean, subtropical South Pacific, and South Indian Oceans all are expected to experience deoxygenation, with O_2 decreases of as much as 17% in the North Pacific by 2100 for the RCP8.5 pathway. However, the tropical Atlantic and tropical Indian Oceans show increasing O_2 concentrations. In the many areas where oxygen is declining, high natural variability makes it difficult to identify anthropogenically forced trends.[128]

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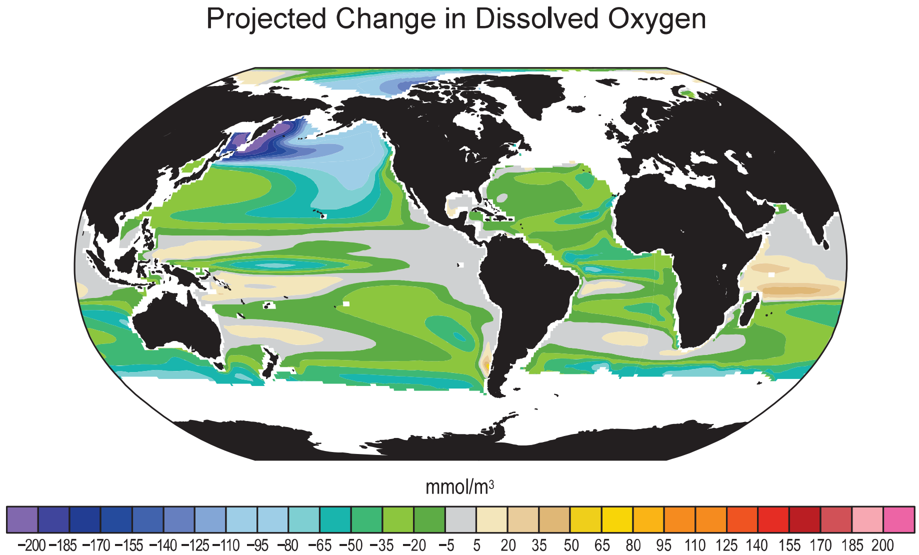

Figure 13.6: Predicted change in dissolved oxygen on the σ_θ = 26.5 (average depth of approximately 290 m) potential density surface, between the 1981–2000 and 2081–2100, based on the Community Earth System Models–Large Ensemble Experiments (Figure source: redrawn from Long et al. 2016[128]).

Regional Models

Regional models are critical because many oxygen drivers are local, influenced by bathymetry, winds, circulation, and fresh water and nutrient inputs. Most eastern boundary upwelling areas are predicted to experience intensified upwelling to 2100,[61] although on the West Coast projections for increasing upwelling for the northern California Current occur only north of San Francisco (see Section 13.2.3).

Particularly notable for the western United States, variation in trade winds in the eastern Pacific Ocean can affect nutrient inputs, leading to centennial periods of oxygen decline or oxygen increase distinct from global oxygen decline.[129] Oxygen dynamics in the Eastern Tropical Pacific are highly sensitive to equatorial circulation changes.[130]

Regional modeling also shows that year-to-year variability in precipitation in the central United States affects the nitrate–N flux by the Mississippi River and the extent of hypoxia in the Gulf of Mexico.[131] A host of climate influences linked to warming and increased precipitation are predicted to lower dissolved oxygen in Chesapeake Bay.[132]

13.5 Other Coastal Changes

13.5.1 Sea Level Rise

Sea level is an important variable that affects coastal ecosystems. Global sea level rose rapidly at the end of the last glaciation, as glaciers and the polar ice sheets thinned and melted at their fringes. On average around the globe, sea level is estimated to have risen at rates exceeding 2.5 mm/year between about 8,000 and 6,000 years before present. These rates steadily decreased to less than 2.0 mm/year through about 4,000 years ago and stabilized at less than 0.4 mm/year through the late 1800s. Global sea level rise has accelerated again within the last 100 years, and now averages about 1 to 2 mm/year.[133] See Chapter 12: Sea Level Rise for more thorough analysis of how sea level rise has already and will affect the U.S. coasts.

13.5.2 Wet and Dry Deposition

Dust transported from continental desert regions to the marine environment deposits nutrients such as iron, nitrogen, phosphorus, and trace metals that stimulate growth of phytoplankton and increase marine productivity.[134] U.S. continental and coastal regions experience large dust deposition fluxes originating from the Saharan desert to the East and from Central Asia and China to the Northwest.[135] Changes in drought frequency or intensity resulting from anthropogenically forced climate change, as well as other anthropogenic activities such as agricultural practices and land-use changes may play an important role in the future viability and strength of these dust sources (e.g., Mulitza et al. 2010[136]).

Additionally, oxidized nitrogen, released during high-temperature combustion over land, and reduced nitrogen, released from intensive agriculture, are emitted in high population areas in North America and are carried away and deposited through wet or dry deposition over coastal and open ocean ecosystems via local wind circulation. Wet deposition of pollutants produced in urban areas is known to play an important role in changes of ecosystem structure in coastal and open ocean systems through intermediate changes in the biogeochemistry, for instance in dissolved oxygen or various forms of carbon.[137]

13.5.3 Primary Productivity

Marine phytoplankton represent about half of the global net primary production (NPP) (approximately 50 ± 28 GtC / year), fixing atmospheric CO_2 into a bioavailable form for utilization by higher trophic levels (see also Ch. 2: Physical Drivers of Climate Change).[138, 139] As such, NPP represents a critical component in the role of the oceans in climate feedback. The effect of climate change on primary productivity varies across the coasts depending on local conditions. For instance, nutrients that stimulate phytoplankton growth are impacted by various climate conditions, such as increased stratification which limits the transport of nutrient-rich deep water to the surface, changes in circulation leading to variability in dry and wet deposition of nutrients to coasts, and altered precipitation/evaporation which changes runoff of nutrients from coastal communities. The effect of the multiple physical factors on NPP is complex and leads to model uncertainties.[140] There is considerable variation in model projections for NPP, from estimated decreases or no changes, to the potential increases by 2100.[141, 142, 143] Simulations from nine Earth system models projected total NPP in 2090 to decrease by 2%–16% and export production (that is, particulate flux to the deep ocean) to drop by 7%–18% as compared to 1990 (RCP8.5).[142] More information on phytoplankton species response and associated ecosystem dynamics is needed as any reduction of NPP and the associated export production would have an impact on carbon cycling and marine ecosystems.

13.5.4 Estuaries

Estuaries are critical ecosystems of biological, economic, and social importance in the United States. They are highly dynamic, influenced by the interactions of atmospheric, freshwater, terrestrial, oceanic, and benthic components. Of the 28 national estuarine research reserves in the United States and Puerto Rico, all are being impacted by climate change to varying levels.[144] In particular, sea level rise, saltwater intrusion, and the degree of freshwater discharge influence the forces and processes within these estuaries.[145] Sea level rise and subsidence are leading to drowning of existing salt marshes and/or subsequent changes in the relative area of the marsh plain, if adaptive upslope movement is impeded due to urbanization along shorelines. Several model scenarios indicate a decline in salt marsh habitat quality and an accelerated degradation as the

rate of sea level rise increases in the latter half of the century.[146, 147] The increase in sea level as well as alterations to oceanic and atmospheric circulation can result in extreme wave conditions and storm surges, impacting coastal communities.[144] Additional climate change impacts to the physical and chemical estuarine processes include more extreme sea surface temperatures (higher highs and lower lows compared to the open ocean due to shallower depths and influence from land temperatures), changes in flow rates due to changes in precipitation, and potentially greater extents of salinity intrusion.

TRACEABLE ACCOUNTS

Key Finding 1

The world's oceans have absorbed about 93% of the excess heat caused by greenhouse gas warming since the mid-20th century, making them warmer and altering global and regional climate feedbacks. Ocean heat content has increased at all depths since the 1960s and surface waters have warmed by about 1.3° ± 0.1°F (0.7° ± 0.08°C) per century globally since 1900 to 2016. Under a higher scenario, a global increase in average sea surface temperature of 4.9° ± 1.3°F (2.7° ± 0.7°C) by 2100 is projected, with even higher changes in some U.S. coastal regions. (*Very high confidence*)

Description of evidence base

The key finding and supporting text summarizes the evidence documented in climate science literature, including Rhein et al. 2013.[4] Oceanic warming has been documented in a variety of data sources, most notably the World Ocean Circulation Experiment (WOCE) (http://www.nodc.noaa.gov/woce/wdiu/) and Argo databases (https://www.nodc.noaa.gov/argo/) and Extended Reconstructed Sea Surface Temperature (ERSST) v4 (https://www.ncdc.noaa.gov/data-access/marineocean-data/extended-reconstructed-sea-surface-temperature-ersst-v4). There is particular confidence in calculated warming for the time period since 1971 due to increased spatial and depth coverage and the level of agreement among independent SST observations from satellites, surface drifters and ships, and independent studies using differing analyses, bias corrections, and data sources.[6, 7, 11] Other observations such as the increase in mean sea level rise (see Ch. 12: Sea Level Rise) and reduced Arctic/Antarctic ice sheets (see Ch. 11: Arctic Changes) further confirm the increase in thermal expansion. For the purpose of extending the selected time periods back from 1900 to 2016 and analyzing U.S. regional SSTs, the ERSST version 4 (ERSSTv4)[19] is used. For the centennial time scale changes over 1900–2016, warming trends in all regions are statistically significant with the 95% confidence level. U.S. regional SST warming is similar between calculations using ERSSTv4 in this report and those published by Belkin,[148] suggesting confidence in these findings. The projected increase in SST is based on evidence from the latest generation of Earth System Models (CMIP5).

Major uncertainties

Uncertainties in the magnitude of ocean warming stem from the disparate measurements of ocean temperature over the last century. There is low uncertainty in warming trends of the upper ocean temperature from 0–700 m depth, whereas there is more uncertainty for deeper ocean depths of 700–2,000 m due to the short record of measurements from those areas. Data on warming trends at depths greater than 2,000 m are even more sparse. There are also uncertainties in the timing and reasons for particular decadal and interannual variations in ocean heat content and the contributions that different ocean basins play in the overall ocean heat uptake.

Summary sentence or paragraph that integrates the above information

There is *very high confidence* in measurements that show increases in the ocean heat content and warming of the ocean, based on the agreement of different methods. However, long-term data in total ocean heat uptake in the deep ocean are sparse leading to limited knowledge of the transport of heat between and within ocean basins.

Key Finding 2

The potential slowing of the Atlantic Meridional Overturning Circulation (AMOC; of which the Gulf Stream is one component)—as a result of increasing ocean heat content and freshwater driven buoyancy changes—could have dramatic climate feedbacks as the ocean absorbs less heat and CO_2 from the atmosphere.[51] This slowing would also affect the climates of North America and Europe. Any slowing documented to date cannot be directly tied to anthropogenic forcing primarily due to lack of adequate observational data and to challenges in modeling ocean circulation changes. Under a higher scenario (RCP8.5) in CMIP5 simulations, the AMOC weakens over the 21st century by 12% to 54% (*low confidence*).

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Description of evidence base

Investigations both through direct observations and models since 2013[4] have raised significant concerns about whether there is enough evidence to determine the existence of an overall slowdown in the AMOC. As a result, more robust international observational campaigns are underway currently to measure AMOC circulation. Direct observations have determined a statistically significant slowdown at the 95% confidence level at 26°N (off Florida; see Baringer et al. 2016[149]) but modeling studies constrained with observations cannot attribute this to anthropogenic forcing.[39] The study[29] which seemed to indicate broad-scale slowing has since been discounted due to its heavy reliance on sea surface temperature cooling as proxy for slowdown rather than actual direct observations. Since Rhein et al. 2013,[4] more observations have led to increased statistical confidence in the measurement of the AMOC. Current observation trends indicate the AMOC slowing down at the 95% confidence level at 26°N and 41°N but a more limited in situ estimate at 35°S, shows an increase in the AMOC.[35, 149] There is no one collection spot for AMOC-related data, but the U.S. Climate Variability and Predictability Program (US CLIVAR) has a U.S. AMOC priority focus area and a webpage with relevant data sites (https://usclivar.org/amoc/amoc-time-series).

The IPCC 2013 WG1 projections indicate a high likelihood of AMOC slowdown in the next 100 years, however overall understanding is limited by both a lack of direct observations (which is being remedied) and a lack of model skill to resolve deep ocean dynamics. As a result, this key finding was given an overall assessment of *low confidence*.

Major uncertainties

As noted, uncertainty about the overall trend of the AMOC is high given opposing trends in northern and southern ocean time series observations. Although earth system models do indicate a high likelihood of AMOC slowdown as a result of a warming, climate projections are subject to high uncertainty. This uncertainty stems from intermodel differences, internal variability that is different in each model, uncertainty in stratification changes, and most importantly uncertainty in both future freshwater input at high latitudes as well as the strength of the subpolar gyre circulation.

Summary sentence or paragraph that integrates the above information

The increased focus on direct measurements of the AMOC should lead to a better understanding of 1) how it is changing and its variability by region, and 2) whether those changes are attributable to climate drivers through both model improvements and incorporation of those expanded observations into the models.

Key Finding 3

The world's oceans are currently absorbing more than a quarter of the CO_2 emitted to the atmosphere annually from human activities, making them more acidic (*very high confidence*), with potential detrimental impacts to marine ecosystems. In particular, higher-latitude systems typically have a lower buffering capacity against pH change, exhibiting seasonally corrosive conditions sooner than low-latitude systems. Acidification is regionally increasing along U.S. coastal systems as a result of upwelling (for example, in the Pacific Northwest) (*high confidence*), changes in freshwater inputs (for example, in the Gulf of Maine) (*medium confidence*), and nutrient input (for example, in agricultural watersheds and urbanized estuaries) (*high confidence*). The rate of acidification is unparalleled in at least the past 66 million years (*medium confidence*). Under the higher scenario (RCP8.5), the global average surface ocean acidity is projected to increase by 100% to 150% (*high confidence*).

Description of evidence base

Evidence on the magnitude of the ocean sink is obtained from multiple biogeochemical and transport ocean models and two observation-based estimates from the 1990s for the uptake of the anthropogenic CO_2. Estimates of the carbonate system (DIC and alkalinity) were based on multiple survey cruises in the global ocean in the 1990s (WOCE – now GO-SHIP, JGOFS). Coastal carbon and acidification surveys have been executed along the U.S. coastal large marine ecosystem since at least 2007, documenting significantly elevated pCO_2 and low pH conditions relative to oce-

anic waters. The data are available from the National Centers for Environmental Information (https://www.ncei.noaa.gov/). Other sources of biogeochemical bottle data can be found from HOT-DOGS ALOHA (http://hahana.soest.hawaii.edu/hot/hot-dogs) or CCHDO (https://cchdo.ucsd.edu/). Rates of change associated with the Palaeocene-Eocene Thermal Maximum (PETM, 56 million years ago) were derived using stable carbon and oxygen isotope records preserved in the sedimentary record from the New Jersey shelf using time series analysis and carbon cycle–climate modelling. This evidence supports a carbon release during the onset of the PETM over no less than 4,000 years, yielding a maximum sustained carbon release rate of less than 1.1 GtC per year.[86] The projected increase in global surface ocean acidity is based on evidence from ten of the latest generation earth system models which include six distinct biogeochemical models that were included in the latest IPCC AR5 2013.

Major uncertainties

In 2014 the ocean sink was 2.6 ± 0.5 GtC (9.5 $GtCO_2$), equivalent to 26% of the total emissions attributed to fossil fuel use and land use changes.[67] Estimates of the PETM ocean acidification event evidenced in the geological record remains a matter of some debate within the community. Evidence for the 1.1 GtC per year cited by Zeebe et al.,[86] could be biased as a result of brief pulses of carbon input above average rates of emissions were they to transpire over timescales $\lesssim$ 40 years.

Summary sentence or paragraph that integrates the above information

There is *very high confidence* in evidence that the oceans absorb about a quarter of the carbon dioxide emitted in the atmosphere and hence become more acidic. The magnitude of the ocean carbon sink is known at a *high confidence* level because it is estimated using a series of disparate data sources and analysis methods, while the magnitude of the interannual variability is based only on model studies. There is *medium confidence* that the current rate of climate acidification is unprecedented in the past 66 million years. There is also *high confidence* that oceanic pH will continue to decrease.

Key Finding 4

Increasing sea surface temperatures, rising sea levels, and changing patterns of precipitation, winds, nutrients, and ocean circulation are contributing to overall declining oxygen concentrations at intermediate depths in various ocean locations and in many coastal areas. Over the last half century, major oxygen losses have occurred in inland seas, estuaries, and in the coastal and open ocean (*high confidence*). Ocean oxygen levels are projected to decrease by as much as 3.5% under the higher scenario (RCP8.5) by 2100 relative to preindustrial values (*high confidence*).

Description of evidence base

The key finding and supporting text summarizes the evidence documented in climate science literature including Rhein et al. 2013,[4] Bopp et al. 2013,[24] and Schmidtko et al. 2017.[150] Evidence arises from extensive global measurements of the WOCE after 1989 and individual profiles before that.[94] The first basin-wide dissolved oxygen surveys were performed in the 1920s.[150] The confidence level is based on globally integrated O_2 distributions in a variety of ocean models. Although the global mean exhibits low interannual variability, regional contrasts are large.

Major uncertainties

Uncertainties (as estimated from the intermodel spread) in the global mean are moderate mainly because ocean oxygen content exhibits low interannual variability when globally averaged. Uncertainties in long-term decreases of the global averaged oxygen concentration amount to 25% in the upper 1,000 m for the 1970–1992 period and 28% for the 1993–2003 period. Remaining uncertainties relate to regional variability driven by mesoscale eddies and intrinsic climate variability such as ENSO.

Summary sentence or paragraph that integrates the above information

Major ocean deoxygenation is taking place in bodies of water inland, at estuaries, and in the coastal and the open ocean (*high confidence*). Regionally, the phenomenon is exacerbated by local changes in weather, ocean circulation, and continental inputs to the oceans.

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14
Perspectives on Climate Change Mitigation

KEY FINDINGS

1. Reducing net emissions of CO_2 is necessary to limit near-term climate change and long-term warming. Other greenhouse gases (for example, methane) and black carbon aerosols exert stronger warming effects than CO_2 on a per ton basis, but they do not persist as long in the atmosphere; therefore, mitigation of non-CO_2 species contributes substantially to near-term cooling benefits but cannot be relied upon for ultimate stabilization goals. (*Very high confidence*)

2. Stabilizing global mean temperature to less than 3.6°F (2°C) above preindustrial levels requires substantial reductions in net global CO_2 emissions prior to 2040 relative to present-day values and likely requires net emissions to become zero or possibly negative later in the century. After accounting for the temperature effects of non-CO_2 species, cumulative global CO_2 emissions must stay below about 800 GtC in order to provide a two-thirds likelihood of preventing 3.6°F (2°C) of warming. Given estimated cumulative emissions since 1870, no more than approximately 230 GtC may be emitted in the future to remain under this temperature threshold. Assuming global emissions are equal to or greater than those consistent with the RCP4.5 scenario, this cumulative carbon threshold would be exceeded in approximately two decades. (*High confidence*)

3. Achieving global greenhouse gas emissions reductions before 2030 consistent with targets and actions announced by governments in the lead up to the 2015 Paris climate conference would hold open the possibility of meeting the long-term temperature goal of limiting global warming to 3.6°F (2°C) above preindustrial levels, whereas there would be virtually no chance if net global emissions followed a pathway well above those implied by country announcements. Actions in the announcements are, by themselves, insufficient to meet a 3.6°F (2°C) goal; the likelihood of achieving that goal depends strongly on the magnitude of global emissions reductions after 2030. (*High confidence*)

4. Further assessments of the technical feasibilities, costs, risks, co-benefits, and governance challenges of climate intervention or geoengineering strategies, which are as yet unproven at scale, are a necessary step before judgments about the benefits and risks of these approaches can be made with high confidence. (*High confidence*)

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Introduction

This chapter provides scientific context for key issues regarding the long-term mitigation of climate change. As such, this chapter first addresses the science underlying the timing of when and how CO_2 and other greenhouse gas (GHG) mitigation activities that occur in the present affect the climate of the future. When do we see the benefits of a GHG emission reduction activity? Chapter 4: Projections provides further context for this topic. Relatedly, the present chapter discusses the significance of the relationship between net cumulative CO_2 emissions and eventual global warming levels. The chapter reviews recent analyses of global emissions pathways associated with preventing 3.6°F (2°C) or 2.7°F (1.5°C) of warming relative to preindustrial times. And finally, this chapter briefly reviews the status of climate intervention proposals and how these types of mitigation actions could possibly play a role in avoiding future climate change.

14.1 The Timing of Benefits from Mitigation Actions

14.1.1 Lifetime of Greenhouse Gases and Inherent Delays in the Climate System

Carbon dioxide (CO_2) concentrations in the atmosphere are directly affected by human activities in the form of CO_2 emissions. Atmospheric CO_2 concentrations adjust to human emissions of CO_2 over long time scales, spanning from decades to millennia.[1, 2] The IPCC estimated that 15% to 40% of CO_2 emitted until 2100 will remain in the atmosphere longer than 1,000 years.[1] The persistence of warming is longer than the atmospheric lifetime of CO_2 and other GHGs, owing in large part to the thermal inertia of the ocean.[3] Climate change resulting from anthropogenic CO_2 emissions, and any associated risks to the environment, human health and society, are thus essentially irreversible on human timescales.[4] The world is committed to some degree of irreversible warming and associated climate change resulting from emissions to date.

The long lifetime in the atmosphere of CO_2[2] and some other key GHGs, coupled with the time lag in the response of the climate system to atmospheric forcing,[5] has timing implications for the benefits (i.e., avoided warming or risk) of mitigation actions. Large reductions in emissions of the long-lived GHGs are estimated to have modest temperature effects in the *near term* (e.g., over one to two decades) because total atmospheric concentration levels require long periods to adjust,[6] but are necessary in the *long term* to achieve any objective of preventing warming of any desired magnitude. Near-term projections of global mean surface temperature are therefore not strongly influenced by changes in near-term emissions but rather dominated by natural variability, the Earth system response to past and current GHG emissions, and by model spread (i.e., the different climate outcomes associated with different models using the same emissions pathway).[7] Long-term projections of global surface temperature (after mid-century), on the other hand, show that the choice of global emissions pathway, and thus the long-term mitigation pathway the world chooses, is the dominant source of future uncertainty in climate outcomes.[3, 8]

Some studies have nevertheless shown the potential for some near-term benefits of mitigation. For example, one study found that, even at the regional scale, heat waves would already be significantly more severe by the 2030s in a non-mitigation scenario compared to a moderate mitigation scenario.[9] The mitigation of non-CO_2 GHGs with short atmospheric lifetimes (such as methane, some hydrofluorocarbons [HFCs], and ozone) and black carbon (an aerosol that absorbs solar radiation; see Ch. 2: Physical Drivers of Climate Change), collectively referred to as short-lived climate pollutants

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(SLCPs), has been highlighted as a particular way to achieve more rapid climate benefits (e.g., Zaelke and Borgford-Parnell 2015[10]). SLCPs are substances that not only have an atmospheric lifetime shorter (for example, weeks to a decade) than CO_2 but also exert a stronger radiative forcing (and hence temperature effect) compared to CO_2 on a per ton basis.[11] For these reasons, mitigation of SLCP emissions produces more rapid radiative responses. In the case of black carbon, with an atmospheric lifetime of a few days to weeks,[12] emissions (and therefore reductions of those emissions) produce strong regional effects. Mitigation of black carbon and methane also generate direct health co-benefits.[13, 14] Reductions and / or avoidances of SLCP emissions could be a significant contribution to staying at or below a 3.6°F (2°C) increase or any other chosen global mean temperature increase.[15, 16, 17, 18] The recent Kigali Amendment to the Montreal Protocol seeks to phase down global HFC production and consumption in order to avoid substantial GHG emissions in coming decades. Stringent and continuous SLCP mitigation could potentially increase allowable CO_2 budgets for avoiding warming beyond any desired future level, by up to 25% under certain scenarios.[18] However, given that economic and technological factors tend to couple CO_2 and many SLCP emissions to varying degrees, significant SLCP emissions reductions would be a co-benefit of CO_2 mitigation.

14.1.2 Stock and Stabilization: Cumulative CO_2 and the Role of Other Greenhouse Gases

Net cumulative CO_2 emissions in the industrial era will largely determine long-term, global mean temperature change. A robust feature of model climate change simulations is a nearly linear relationship between cumulative CO_2 emissions and global mean temperature increases, irrespective of the details and exact timing of the emissions pathway (see Figure 14.1; see also Ch. 4: Projections). Limiting and stabilizing warming to any level implies that there is a physical upper limit to the cumulative amount of CO_2 that can be added to the atmosphere.[3] Eventually stabilizing the global temperature requires CO_2 emissions to approach zero.[19] Thus, for a 3.6°F (2°C) or any desired global mean warming goal, an estimated range of cumulative CO_2 emissions from the current period onward can be calculated. The key sources of uncertainty for any compatible, forward looking CO_2 budget associated with a given future warming objective include the climate sensitivity, the response of the carbon cycle including feedbacks (for example, the release of GHGs from permafrost thaw), the amount of past CO_2 emissions, and the influence of past and future non-CO_2 species.[3, 19] Increasing the probability that any given temperature goal will be reached therefore implies tighter constraints on cumulative CO_2 emissions. Relatedly, for any given cumulative CO_2 budget, higher emissions in the near term imply the need for steeper reductions in the long term.